

# The Soil Freezing Characteristic: Its Measurement and Similarity to the Soil Moisture Characteristic

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## ABSTRACT

A soil freezing characteristic (SFC) represents the relationship between the quantity and the energy status of liquid water in frozen soil. The SFC is the analogue to the soil moisture characteristic (SMC) and is essential to modeling the transport of water, heat, and solutes in frozen soil. This paper presents a new, automated technique to measure an SFC in situ, for which there has previously been no method. Liquid water content in frozen soil was measured with time domain reflectometry. The corresponding energy status was inferred from accurate soil temperature measurements with a generalized form of the Clapeyron equation. Since both SFC and SMC describe water retention properties in soil, their similarity was investigated. The SMC and SFC agreed to within 1% moisture content across a wide range of matric potentials. Determination of the SMC is reliable at high matric potentials but becomes increasingly inaccurate and time consuming as soil dries. By contrast, the SFC determination becomes more accurate and rapid at lower matric potentials. We thus propose that water retention properties at high matric potentials are best obtained from draining and at low matric potentials from freezing.

WHEN THE TEMPERATURE DROPS below the freezing point of a soil, a portion of the soil water changes phase and freezes in situ. Water that remains unfrozen is present in adsorbed films around particles, in crevices between particles, and in pores of sufficiently small diameter. As temperature declines further, more water will freeze, leaving the remaining unfrozen water in progressively thinner adsorbed films and smaller pores and crevices. Soil water can remain unfrozen at temperatures well below the freezing point because matric forces exerted by soil on the water lower the energy status of the water, which, consequently, lowers the freezing point of the water. Thus, in frozen soil (in this study, the terms frozen and freezing indicate the presence of ice), where liquid water and ice coexist, water potential is strongly dependent on temperature; their relation can in fact be derived from thermodynamics. The relationship between water potential and (liquid) water content in frozen soil is called the SFC, analogous to the definition of an SMC for unfrozen soil (Miller, 1966).

The presence of a liquid water phase below the freezing point has significant consequences for the transport of water, solutes, and heat in frozen soil. Water movement in frozen soil is evident from phenomena such as frost heave where water migrates through a frozen layer towards a growing ice lens (Miller, 1980) and the infiltration of large amounts of snowmelt into frozen soil as reported by Baker and Spaans (1994). Thus, knowledge of the SFC is of primary importance when studying transport processes in frozen soil. However, there has

previously been no method for in situ determination of the SFC, mainly because of inadequate means to accurately discriminate between liquid water and ice.

Since both SFC and SMC describe moisture retention properties of soils, it has been hypothesized that the two curves should superimpose if plotted in an appropriate way (Koopmans and Miller, 1966). This would allow the construction of soil moisture retention curves with data from drying and wetting cycles supplemented by freezing and thawing data at low matric potentials where drying and wetting data are difficult to obtain. The objectives of this study were to develop a new technique to determine SFC in the field or laboratory on intact soil at any degree of saturation and to investigate the similarity between the SFC and the SMC for a silt loam.

## MATRIC POTENTIAL IN FROZEN SOIL

Water potential in frozen soil ( $\Psi_L$ , in J kg<sup>-1</sup>), where liquid water and ice coexist, can be calculated from a generalized form of the Clapeyron equation:

$$d\Psi_L = (\lambda/T)dT + d\Psi_i \quad [1]$$

where  $\Psi_i$  is the ice potential (J kg<sup>-1</sup>),  $T$  is temperature (K), and  $\lambda$  is the latent heat of fusion of water (333.7 kJ kg<sup>-1</sup> at 273.15 K). Subscripts L and i refer to liquid water and ice, respectively. More detail about the derivation of Eq. [1] is given by Lewis and Randall (1961, p. 151) and Spaans (1994). The important implication of Eq. [1] is that once water and ice coexist, the water potential is determined only by temperature and ice pressure. If ice pressure is known, a temperature measurement is therefore sufficient to calculate the water potential. Note that  $\Psi$  is the total potential, which is the sum of osmotic potential ( $\Pi$ ) and matric potential ( $\psi$ ). Since solutes are rejected from the ice phase,  $\Pi_i = 0$ , and thus  $\Psi_i = \psi_i$ .

The value for  $\lambda$  is temperature dependent, but only few data are available (List, 1951, Table 92), for which a best fit yields  $\lambda(T) = -712.38 + 5.545T - 6.28 \times 10^{-3} T^2$ . Choosing pure water at atmospheric pressure and a temperature  $T_0 = 273.15$  K as a reference state, and assuming ice at atmospheric pressure, the integral form of Eq. [1] becomes

$$\Psi_L = -712.38 \ln(T/T_0) + 5.545(T - T_0) - 3.14 \times 10^{-3}(T^2 - T_0^2) - \Pi_L \quad [2]$$

The broad assumption of zero gauge pressure in the ice phase has been questioned under certain conditions (Miller, 1973, 1980), but thus far there is scant evidence against it, except in obvious cases (heaving).

As pore ice forms, solutes are rejected into the unfrozen water layers and become increasingly concentrated. Banin and Anderson (1974) showed that the effects of solutes on the freezing point depression of pore water are generally predictable on the basis of thermodynamic arguments if one takes into account the gradual concentration of solutes in the remaining water due to exclusion from the ice phase. They

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**Abbreviations:** SFC, soil freezing characteristic; SMC, soil moisture characteristic; TDR, time domain reflectometry; PRT, platinum resistance thermometer.

concluded that soil solution generally behaves as an ideal solution and that secondary effects, such as the heat of mixing, solute effects on water structure, or the nonhomogeneous distribution of solute, may be operative but are usually negligible. Williams (1964) extracted water from soil samples at different pressures to examine the distribution of solutes in soil water. He found, for several soil types, that solute concentrations of soil solutions extracted at pressures between 0 and 1.4 MPa were similar, indicating that solutes were homogeneously distributed in soil water retained at  $\psi > -1.4 \text{ kJ kg}^{-1}$ . Consequently, when solute movement during freezing is negligible, the osmotic potential in frozen soil can be estimated from liquid water content, temperature, and osmotic potential of the soil solution prior to freezing.

### LIQUID WATER CONTENT IN FROZEN SOIL

The presence of an ice phase complicates detection of unfrozen water in frozen soil. Commonly used gravimetry and neutron moderation do not discriminate between liquid water and ice and thus yield total water content. Dilatometry has been successfully used (Koopmans and Miller, 1966; Patterson and Smith, 1981) to measure SFC in the laboratory, but this technique requires a water-saturated sample. This is usually accomplished by slurring the sample, thereby destroying the original soil structure. More recently, nuclear magnetic resonance has been employed to estimate the unfrozen water content but only for small, repacked samples (Tice et al., 1982; Black and Tice, 1989), since this technique is unsuited to field research (Oliphant, 1985).

Patterson and Smith (1981) showed that TDR is a promising technique to obtain the volumetric liquid water content ( $\theta_L$ ) in frozen soil, because the permittivity ( $\epsilon$ ) of water is much higher than that of other soil constituents, including ice. Application of TDR requires a calibration equation relating the soil  $\epsilon$  to  $\theta_L$ . Several such equations exist for unfrozen soil and have been applied in some cases without modification to frozen soils (Stein and Kane, 1983; Patterson and Smith, 1981). However, this may not be appropriate, since the  $\epsilon$  for ice is higher than that for air. A recently developed method with a gas dilatometer (Spaans and Baker, 1995) allows independent, accurate calibration of TDR for  $\theta_L$  in frozen, intact soil at any degree of saturation. This calibration, coupled with the automation and multiplexing capabilities of TDR (Baker and Allmaras, 1990) and improved parallel waveguides (Spaans and Baker, 1993), provided the opportunity to unequivocally determine an SFC in the field.

### SOIL FREEZING AND SOIL MOISTURE CHARACTERISTIC SIMILARITY

Drying and wetting phenomena in unfrozen soil are strikingly similar to freezing and thawing phenomena in frozen soil. As soils dry, water is removed and replaced by air leaving the remaining water at an increasingly lower matric potential. The same occurs in soils that freeze, except that liquid water changes phase and becomes ice. The same forces that prevent soil water from draining also prevent it from freezing. This apparent similarity has prompted a hypothesis that the SMC and the SFC contain similar information, and therefore soil water retention properties can be obtained by either drying and wetting or freezing and thawing (Schofield, 1935; Miller, 1966; Williams, 1964).

Water retention in soils may be attributed to interactions between soil and liquid water that involve two kinds of interfacial phenomena. Adsorptive forces act at the water-mineral interfaces, while capillary forces act at water-air, water-ice,

and air-ice interfaces in pore necks or crevices that connect the inner pore body with the exterior (Miller, 1966). Miller (1966) postulated that parallels between the SMC and the SFC are quantitative only for soil water that is held exclusively by either capillary or adsorptive forces. Such conditions are simulated in pure sand or silt and in dispersed clay, respectively. His theorem was confirmed by experiment (Koopmans and Miller, 1966). Most natural soils, however, have attributes of both extremes, and the parallel can only be quantitative insofar as the soil approaches one of the extreme states.

This hypothesis followed from the theory of *similitude* (Miller and Miller, 1956). If the same states of soil moisture content and distribution are achieved by freezing and thawing as by drying and wetting, the two states of water content are said to be *similar*. Accordingly, Koopmans and Miller (1966) insisted that drying and wetting of ice-free soil can only be compared with freezing and thawing of air-free soil. In addition, since both SMC and SFC are hysteretic, the first drying cycle can only be compared with the first freezing cycle, and the same is true for wetting and thawing cycles. Consequently, to superimpose an SMC onto an SFC, we must compare the water potential with respect to air in unfrozen soil on one hand, with the water potential with respect to ice in frozen soil on the other hand, at similar states of water content.

The adsorptive component of the matric potential, which acts on the mineral-water interface, acts only in the liquid phase since it is the only medium in which ions of the double layer exist (Miller, 1980). It is therefore considered immaterial whether ice or air is present on the other side of the water film, thus the reduction of the soil water potential due to adsorptive forces is identical in both frozen and unfrozen states. Capillary forces, however, are proportional to the surface energy ( $\gamma$ ) of the interface, which is higher for water-air interfaces ( $\gamma_{wa}$ ) than for water-ice interfaces ( $\gamma_{wi}$ ). Hence, if the soil moisture is at similar states in the frozen and unfrozen soil with capillary forces dominating, the pressure difference between water and ice must be a factor  $\gamma_{wa}/\gamma_{wi}$  larger than the pressure difference between water and air. By experiment, Koopmans and Miller (1966) found that the value of  $\gamma_{wa}/\gamma_{wi}$  was 2.2.

This quantitative comparison between SMC and SFC offered a unique opportunity to transpose data of moisture retention properties from unfrozen soil, which were either readily available or easily measured, to frozen soil, thus circumventing direct measurement of the SFC (Miller, 1966; Black and Tice, 1989). However, three obstacles arise. First, an ice-free SMC can be compared only with an air-free SFC. But frozen soils in the field are not necessarily saturated since water, ice, and vapor can coexist in equilibrium in the pore space of frozen soils (Miller, 1972; Colbeck, 1982). Second, the quantitative rules to match SMC and SFC are supposedly valid only for colloidal soils or for colloid-free soils. Most soils in the field combine both types of behavior to some degree, and the transition from capillary to adsorbed water is gradual. Finally, determination of the SMC becomes increasingly inaccurate and time consuming as soil dries, so justification for this transformation at lower matric potentials is questionable. The time required to attain equilibrium after changes in pressure during stepwise desorption or adsorption experiments is dependent on the soil hydraulic conductivity, sample height, and the hydraulic contact between sample and porous plate. Since the hydraulic conductivity declines sharply as the soil dries, equilibration times can increase considerably at low water contents. Spaans (1994) calculated that it takes as much as 12 d to change the matric potential from  $-1$  to  $-1.5 \text{ kJ kg}^{-1}$  in a 3-cm-tall sample placed in a Richards apparatus, assuming perfect contact between sample and plate. Campbell (1988)

reported that for SMC determination at matric potentials lower than  $-0.5 \text{ kJ kg}^{-1}$ , the low hydraulic conductivity itself can cause substantial errors but that lack of equilibrium is often the result of poor contact between sample and porous plate. He concluded that much of the pressure plate data are likely to be in error at low water potential. Work by Madsen et al. (1986) showed that the water potential of samples equilibrated on a pressure plate measured with a thermocouple psychrometer was only half of the expected water potential. Apparently, the samples never reached equilibrium at low water potentials although they were only 1 cm high and were equilibrated for at least 34 d.

Fortuitously, while the determination of the SMC is most accurate in wet soils and becomes increasingly inaccurate as the soil dries, the determination of the SFC becomes more accurate once the loosely held water is frozen. When most capillary water has drained or frozen and liquid water is held mainly by adsorptive forces, then water retention in natural soils might resemble that as it occurs in colloidal soils, and similarity between SMC and SFC is established. We hypothesize, therefore, that water retention properties for the wet range can most accurately be obtained from the SMC, whilst those in the dry range can be obtained most accurately from the SFC.

In summary, to obtain an SFC, one must plot the matric potential with the corresponding  $\theta_L$  in frozen soil. Similitude analysis of the SFC and the SMC, however, requires an additional step: the matric potential in frozen soil must be compared with that in unfrozen soil at equal  $\theta_L$ , and the rules that apply depend on whether the water is held by adsorptive forces or in capillaries with ice on the other side of the interface. In the former case,  $\psi_L(\text{unfrozen soil}) = \psi_L(\text{frozen soil})$ , while in the latter case,  $\psi_L(\text{unfrozen soil}) = 2.20 \psi_L(\text{frozen soil})$ .

## MATERIALS AND METHODS

### Soil Freezing Characteristic

Determination of an SFC involves simultaneous measurement of potential and quantity of unfrozen water in frozen soil. These two parameters were measured in the field at the University of Minnesota's Rosemount Experimental Station, located  $\approx 30 \text{ km}$  south of St. Paul, MN ( $44^\circ 42' \text{ N}$ ,  $93^\circ 32' \text{ W}$ , 290 m elevation). The soil is classified as a Waukegan series (fine-silty over sandy or sandy-skeletal, mixed, mesic Typic Hapludoll). The texture is silt loam to a depth of  $\approx 0.7 \text{ m}$ , and the mollic layer typically contains  $>20 \text{ g kg}^{-1}$  organic C.

Our automated TDR system to determine  $\theta_L$  is a modification of the digital system described by Baker and Allmaras (1990). To minimize signal-energy losses, the entire transmission line, including switches, is coaxial with a characteristic impedance of  $50 \Omega$ . The transmission line terminates in a balanced probe, equipped with an SB1:1 balun in the head of the probe where the coaxial line joins the parallel rods (Midwest Special Services<sup>1</sup>, St. Paul, MN). This probe is similar to that described by Spaans and Baker (1993), except that the rod spacing is 3 cm instead of 5 cm, which yields a more evenly distributed electrical field (Knight, 1992).

Probes [length ( $L$ ) = 0.3 m, rod diam. = 3.2 mm] were installed in the field from a single pit at 5-, 10-, 15-, 20-, 31-, 46-, 70-, and 108-cm depth into one pit face and at 7-, 17-, 29-, and 46-cm depth into the opposite pit face. The probes were pushed horizontally into the pit faces (parallel to the isotherms in the soil) with the aid of a wooden jig to

keep the rods parallel and to avoid air gaps around the rods. Approximately 0.5 m of the coaxial cable was coiled behind the probe at the same depth before it was brought to the surface, to minimize heat conduction through the cable to the area of measurement.

Above the surface, the coaxial cables were connected to eight-position coaxial switches (Model 50S-608, JFW, Indianapolis, IN), housed in a weather-proof enclosure. The appropriate output port on these switches was closed by providing 12 VDC (70 mA) across one of the pins of a connector on the switch. The coaxial input of each eight-position switch was connected to a central two-position coaxial switch (JFW Model 50S-597) by RG-8/U cable (Belden 8214), which is stout, low-loss  $50 \Omega$  coaxial cable. The input of the two-position switch was, in turn, connected to a TDR (Tektronix 1502c, Beaverton, OR), which was powered by a 12 VDC power supply. The TDR, the two-position switch, and the power supply were placed in an insulated weather-proof enclosure; the heat dissipating from the power supply kept the temperature in the enclosure above freezing.

On the edge of the field, 60 m south of the pit, we had a heated trailer to shelter a personal computer and to thaw out scientists. Serial communication between the computer and the TDR was established by equipping the TDR with an SP232 module (Tektronix), to convert the parallel port of the TDR to a serial port. On both ends of the serial cable, optical isolators (B&B Electronics Model 232OP4, Ottawa, IL) were inserted to protect both computer and TDR from lightning and damage due to ground potential differences. A parallel I/O board (Metabyte Corp., Taunton, MA) inside the computer allowed computer control of a relay board that routed 12 VDC from a power supply in the trailer to the appropriate pin on the coaxial switches.

Traces were acquired from all TDR probes every half hour around the clock and were analyzed for apparent rod length ( $L_a$ ). The trace-analysis routine of Baker and Allmaras (1990) was modified, since traces from probes with an SB1:1 balun differ from those obtained with 1:4 baluns, due to a different output impedance (Spaans and Baker, 1993). In frozen or very dry soil, the point where the signal enters the soil from the probes with an SB1:1 is not readily apparent, so the initial reflection was found by detecting the point where the signal leaves the balun, which is a clear mark on the trace. The travel time from this mark to the point where the signal enters the soil is 0.23 ns for all probes and is independent of the soil conditions since the balun is shielded and embedded in epoxy. This travel time adds 0.069 m to  $L_a$  which must be subtracted to yield the true  $L_a$ .

Spaans and Baker (1995) found that there is no unique calibration for TDR in frozen soil, but rather a family of calibration curves, each corresponding to a different total water content. Therefore, intact soil samples for a TDR calibration were taken shortly before the soil started to freeze to determine a calibration curve corresponding to the total water content in the field at the depths of interest, which yielded

$$\theta_L = 0.1161L_a/L - 0.1747 \quad [3]$$

Matric potential was calculated from Eq. [2]. Soil temperatures were measured with thermistors (YSI no. 44004, Yellow Springs, OH), installed at the same depth as the TDR probes. On either side of each probe, one thermistor was inserted 15 cm into the soil at a distance of  $\approx 3 \text{ cm}$  from the probe, to stay outside the sphere of influence of the TDR signal (Baker and Lascano, 1989). These thermistors were multiplexed (Campbell Scientific Model AM32, Logan, UT) and their resistances measured in a six-wire Wheatstone bridge with a

<sup>1</sup> Mention of manufacturers is for the convenience of the reader only and implies no endorsement on the part of the authors, University of Minnesota, or USDA.

datalogger (Campbell Scientific 21X) simultaneous with the computer-based TDR measurement.

Prior to installation, all thermistors were calibrated against a platinum resistance thermometer (secondary temperature standard PRT, accuracy 0.005°C; Minco Model S7929, Minneapolis, MN) in the laboratory. Thermistors were attached to the PRT, submerged in an insulated container filled with an ethylene-glycol solution, and placed on a magnetic stirrer. The temperature of the fluid was cycled in 36 steps between -15 and +10°C by means of a constant temperature bath. The resistances of each individual thermistor were then fitted to the corresponding temperatures from the PRT with the Steinhart-Hart equation:

$$1/T = a + b \ln R + c(\ln R)^3 \quad [4]$$

where  $R$  is the resistance ( $\Omega$ ) of the thermistor,  $T$  is the temperature (K) obtained from the PRT, and  $a$ ,  $b$ , and  $c$  are fitting coefficients unique for every thermistor. Residuals for each fit were typically  $<0.02^\circ\text{C}$ .

Osmotic potential ( $\text{kJ kg}^{-1}$ ) of the unfrozen water was estimated from electrical conductivity measurements on saturated paste extracts ( $\sigma_{\text{sp}}$ ) (Rhoades, 1986; Banin and Anderson, 1974; Cary et al., 1979; Williams, 1964):

$$\Pi_L = -39\sigma_{\text{sp}}(W_{\text{sp}}\rho_b)/(\theta_L\rho_L^2)(T/298) \quad [5]$$

assuming that none of the salts would precipitate when water is frozen out of solution. At very low  $\theta_L$  this assumption may not hold, but by then, the temperature would have dropped so far that the contribution of osmotic potential to the total water potential is insignificant compared with the matric component. The soil was sampled in 5-cm depth increments shortly before the soil froze to determine  $\sigma_{\text{sp}}$  ( $\text{mS cm}^{-1}$ ) and soil bulk density ( $\rho_b$  in  $\text{kg m}^{-3}$ ). The parameter  $W_{\text{sp}}$  is the wetness (typically  $0.5 \text{ kg kg}^{-1}$ ), and 298 is the temperature at which  $\sigma_{\text{sp}}$  was determined;  $T$  is soil temperature (K);  $\rho_L$  is the density of water ( $\text{kg m}^{-3}$ ); and  $\theta_L$  was obtained from the TDR measurements.

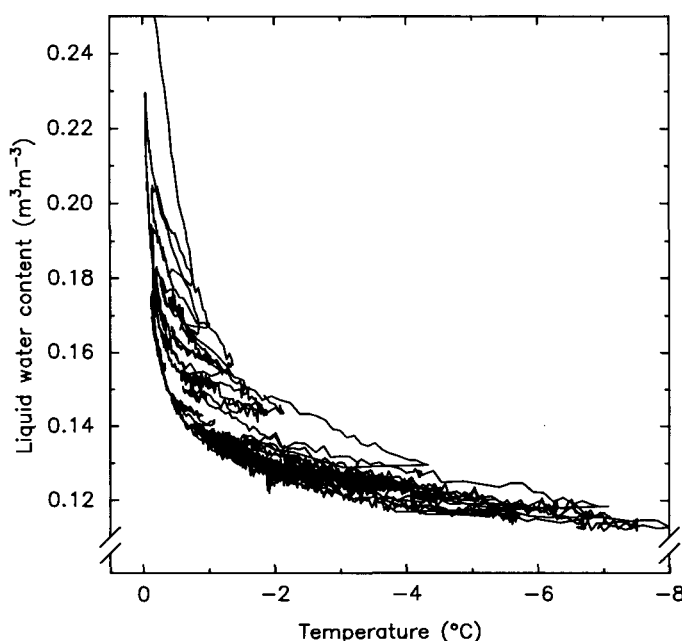


Fig. 1. Liquid water content at 5-cm depth as a function of temperature during the entire winter. Scanning curves demonstrate hysteresis. Data on the upper boundary of the envelope represent the first freezing cycle of the SFC.

## Soil Moisture Characteristic

We decided against in situ determination of the SMC primarily because of the difficulty of measuring matric potential with sufficient accuracy in the field across the entire range of interest. In addition, in recent years, the soil in our field has not dried far enough to measure a complete SMC in situ. Thus, undisturbed core samples (8.5 cm diam., 7 cm height) were taken close to where SFCs were measured and were placed in Tempe cells (Soil Moisture, Santa Barbara, CA) for SMC determination in the laboratory. Two samples were taken from the bottom of the Ap (14–21 cm) and two from the top of the Bt1 (30–37 cm) horizon. In the laboratory, the samples were slowly saturated and subsequently drained by a hanging water column to a matric potential of  $-20 \text{ J kg}^{-1}$  and from  $-20$  to  $-100 \text{ J kg}^{-1}$  by applying air pressure to the sample. To obtain matric potentials from  $-100$  to  $-1500 \text{ J kg}^{-1}$ , the samples were transferred from the Tempe cells to a porous plate in a pressure chamber.

After a step change in pressure, either outflow or weight was periodically recorded to check for equilibrium. Once the pressure chamber was employed, the total weight of the porous plate with the samples intact was recorded to minimize disturbance of the contact between sample and plate; only at equilibrium were the samples individually weighed. At the end of the experiment, all samples were oven dried at  $105^\circ\text{C}$  for 48 h, and moisture content was calculated from the cumulative weight losses.

## RESULTS AND DISCUSSION

### Soil Freezing Characteristics

Data for the SFCs were obtained during the winter of 1992–1993. The surface soil started freezing around 1 Dec. and thawed in late March. Freezing reached a maximum depth of 70 cm. The soil typically thaws from both top and bottom, so the 46-cm depth did not thaw until mid-April.

Figure 1 shows measurements of soil  $T$  versus  $\theta_L$  at 5 cm during the entire winter. Considerable hysteresis in the SFC is apparent from the scanning curves and is, at least qualitatively, similar to hysteresis in drying and wetting curves observed in SMCs. In this paper, however, we will focus on the first freezing cycle of the SFC. All paired data of  $T$  and  $\theta_L$  were obtained within 5 min of each other, and only those data pairs were extracted where the temperature was lower than any previous temperature. Thus, when a transient warming trend occurred, no data were extracted until the temperature had dropped again to a value below that prior to the warming trend. Data acquisition was interrupted occasionally by computer problems and power failures. The most complete data sets were obtained from the 29- and 46-cm depths since they froze after the system was fully operational.

The SFCs for the 29- and 46-cm depths are shown in Fig. 2, where  $\psi_L$  was calculated from Eq. [2] and [5]. Liquid water contents are shown on a gravimetric scale ( $W_L$ ) to account for differences in bulk densities, which were  $1.38 \text{ Mg m}^{-3}$  for the 29 cm, and  $1.34 \text{ Mg m}^{-3}$  for the 46-cm depth. The data in Fig. 2 have neither been smoothed nor filtered; nevertheless, they show remarkably little scatter. Figures 2B and 2C are duplicates since SFC was observed in two opposite sides of the pit.

Data in the flat region of the SFCs (at  $\psi_L > -50 \text{ J}$

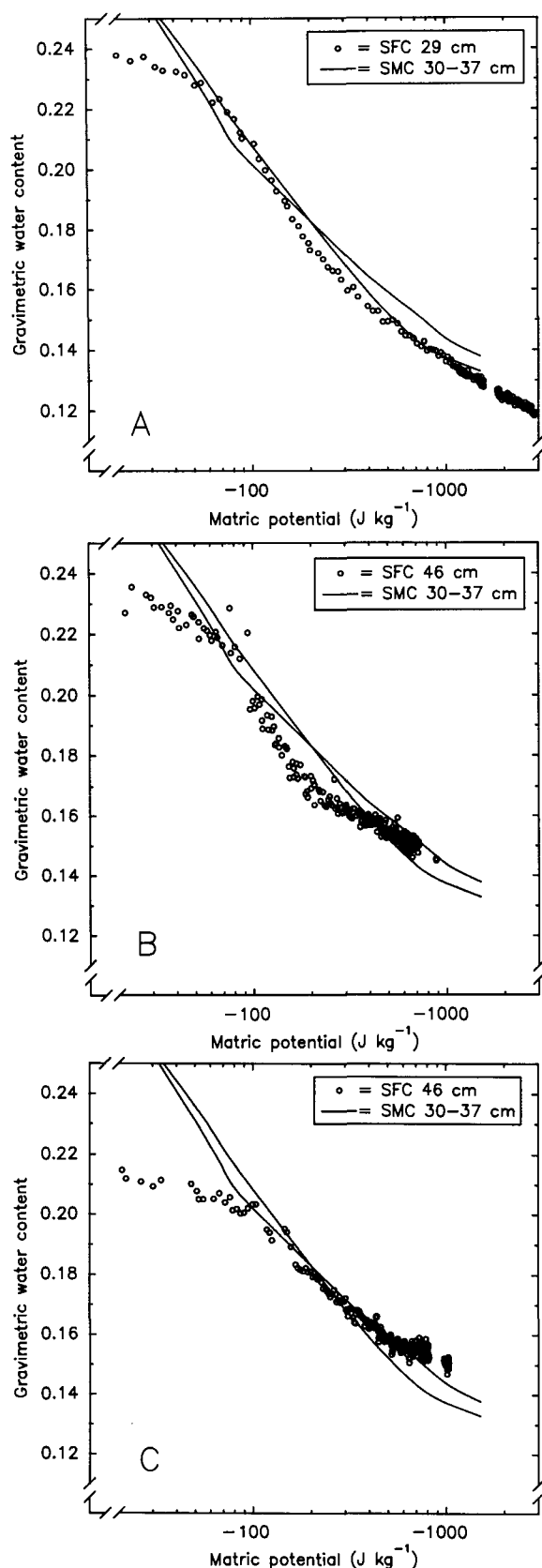


Fig. 2. (A) Soil freezing characteristics (SFC) for the 29-cm depth and (B and C) the 46-cm depth from two opposite sides of the pit, and the soil moisture characteristic (SMC) for 30- to 37-cm depth (duplicates). Data at high matric potential in the flat region of the SFC represent unfrozen soil immediately prior to freezing.

$\text{kg}^{-1}$  in Fig. 2A and B, and  $\psi_L > -100 \text{ J kg}^{-1}$  in Fig. 2C) are actually not part of the SFC because they represent unfrozen conditions, but they were included to demonstrate the dynamics of the water regime in the soil immediately prior to freezing. As the freezing front penetrates, the unfrozen soil ahead of it will dry due to water migration towards the ice front (Dirksen and Miller, 1966). This is evident from the slight drop in  $W_L$  in Fig. 2 prior to freezing. The reduced matric potential in the unsaturated soil caused a freezing point depression in accordance with Eq. [2], i.e., the soil was too dry to freeze at first. Note that  $\psi_L$  on the  $x$  axis in Fig. 2 is calculated from Eq. [2], so it does not represent the actual matric potential until liquid water and ice coexist.

With TDR-based measurements of  $\theta_L$  in the field, one cannot tell exactly when nucleation is initiated because the system is not closed. Immediately below the ice front, strong matric potential gradients exist across relatively short distances, so rapid dewatering might occur, which gradually ceases when freezing commences. This is illustrated in Fig. 3, which shows a time series of the thermistor and TDR measurements at 29-cm depth. A transitional decline in  $\theta_L$  is apparent before  $\theta_L$  drops sharply due to freezing, which occurs at  $\theta_L \approx 0.31$ .

Calculating water potential in frozen soil from temperature requires highly accurate temperature measurements. Thermistors have an advantage over thermocouples because of their large response to minute temperature differences ( $380 \Omega \text{ K}^{-1}$  at  $T_0$ ), but their water resistance and stability is a concern. If moisture penetrates the semi-conductor material of the thermistor, its response to temperature is altered irreversibly. Thermocouples, on the other hand, are much less sensitive to temperature and require considerable care to obtain a reliable reference temperature measurement and to avoid thermal gradients along the panels of multiplexing and measurement devices.

### Soil Freezing and Soil Moisture Characteristic Similarity

The first drainage cycles of the SMC for the Ap and the Bt1 horizons are shown in Fig. 4. Pore saturation was only 85 to 90% at the start of the drainage cycle, which is typical for this procedure (Klute, 1986). The Bt1 horizon retains less water at lower  $\psi$ , probably because of its lower organic matter content compared with the Ap horizon. The bulk densities of the two samples from the Bt1 horizon are similar; those from the Ap horizon differ by almost 10%, yet they retain the same amount of water per volume of soil close to saturation. As the largest pores have drained, however, the sample with the higher bulk density retains more water per volume of soil. Beyond  $\approx -20 \text{ J kg}^{-1}$ , that difference disappears when the amount of water is expressed per unit mass of soil, which is evidence that we are dealing with adsorbed water. This may be trivial for the SMC, but it is important for the analysis of similarity between SMC and SFC, because different rules apply for capillary and adsorbed water.

The water contents of the 29- and 46-cm depths before they froze corresponded to approximately  $\psi < -50 \text{ J kg}^{-1}$  according to the SMC. This indicates that the soil

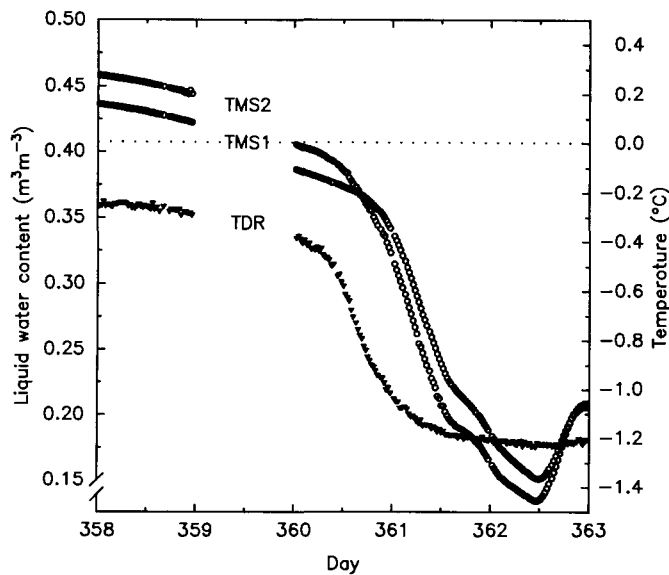


Fig. 3. Soil freezing at 29-cm depth. Drop in liquid water content prior to freezing demonstrates water migration towards the frozen soil above. Freezing occurred on Day 360. Data gap is due to power failure in the trailer. TMS refers to thermistors located on either side of the time domain reflectometry (TDR) probe.

water present as freezing commenced was mainly adsorbed water, which has two implications. First, at equal moisture contents,  $\psi$  in frozen soil is equal to that in unfrozen soil without the need to adjust for the differences in surface tension between ice-water and air-water interfaces. Second, water content should be presented on a gravimetric scale to account for differences in  $\rho_b$  when comparing an SFC and an SMC.

The SMC from the 30- to 37-cm depth is compared with the SFC measured in the field at 29- and 46-cm depth in Fig. 2. The agreement between the SMC and the SFC is excellent, with  $<0.01 \text{ g g}^{-1}$  difference in  $W$  at any  $\psi$  across a wide range of matric potentials. The temperature at 46 cm did not drop far enough for a complete comparison to be made between the two characteristics.

In regions not blessed with such a bracing climate, SFCs can be determined on intact soil cores fitted with thermistors and waveguides, then frozen in steps in the laboratory. However, supercooling and temperature stability limit laboratory determination of SFCs. In the absence of ice, soil water can be supercooled several degrees below its freezing point, which is common in the laboratory but rarely happens in the field (Anderson and Morgenstern, 1973; Miller, 1980; personal observations). Once soil water is supercooled, a small disturbance may induce ice nucleation after which the ice grows until it is in equilibrium with the liquid soil water at that temperature (i.e., satisfying Eq. [1]). Thus in that respect, supercooling prohibits measurement of the first freezing cycle of the SFC at high values of  $\psi_L$ . This is, however, not a problem for the cold end of the SFC. Temperature control and stability require utmost attention.

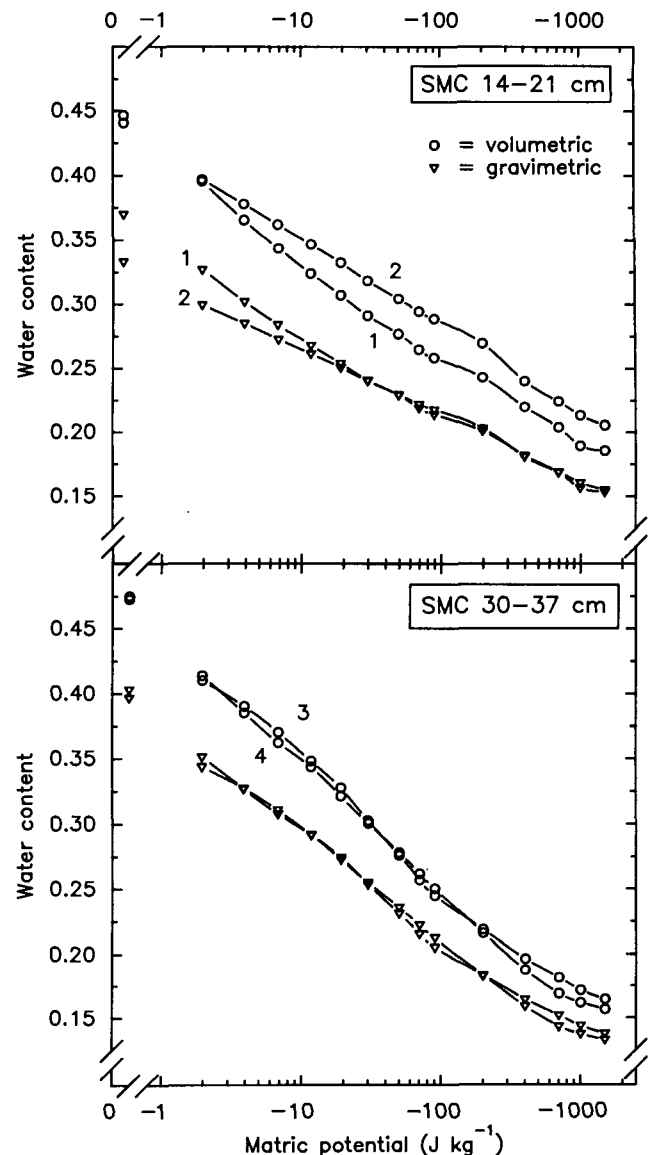


Fig. 4. Soil moisture characteristics (SMC) for the Ap horizon (14-21 cm) and the Bt1 horizon (30-37 cm). Both graphs show water content on a gravimetric and a volumetric scale. Bulk densities for Samples 1 through 4 are  $1.21$ ,  $1.32$ ,  $1.19$ , and  $1.18 \text{ Mg m}^{-3}$ , respectively.

## SUMMARY AND CONCLUSIONS

Soil moisture characteristics compared favorably with soil freezing characteristics when (i) ice was assumed at atmospheric pressure, (ii) water content was expressed gravimetrically, and (iii) water-ice interfacial forces were neglected. Soil morphological observations in the field revealed that ice was present in large crystals in worm and root channels and between peds. Such ice bodies must form at ambient pressure in an unconfined sample (Koopmans and Miller, 1966). The second and third assumptions are interrelated, since the soil was unsaturated to the point where adsorptive forces dominated water retention. Therefore, when water retention is compared between samples of different bulk densities, the water content should be normalized to the mass of

soil present. Since adsorptive forces act only on the liquid phase, it is immaterial whether ice or air is present at the other side of the soil-water interface, so ice-water interfacial energies can, in this case, be neglected.

From the data presented in this study, nothing can be said about the comparison between the SMC and the SFC at higher matric potentials. Obtaining reliable data on the SFC at high  $\theta_L$  is difficult due to the following practical limitations. Uncertainties in the temperature measurement are very critical close to the freezing point, where much of the soil water freezes across a relatively narrow temperature range. In the surface soil, spatial variation of temperature is large, so the distance between the thermistors and the TDR probes is of concern. The dampened temperature fluctuations and slow phase changes in the subsoil are both favorable for accurate SFC determinations. These conditions are also favorable for dewatering soil before it freezes, so again it is difficult to measure the SFC at high  $\theta_L$ . Finally, at temperatures immediately below the freezing point, the contribution of the osmotic potential ( $-70 \text{ J kg}^{-1}$  for both depths in this study) to the total water potential is relatively large, so uncertainties in the estimation of the osmotic potential are significant. In addition, freezing-induced water migration adversely affects the accuracy of Eq. [5]. An attractive alternative would be to estimate in situ osmotic potential from TDR-based bulk electrical conductivity measurements.

A complete SFC is required to understand heat, water, and solute transport in frozen soil. The automated TDR system in concert with thermistors and an independent TDR calibration provided an extensive set of field data on liquid water content and temperature in frozen soil. For reasons described earlier, field measurements of the SFC at high  $\psi_L$  are difficult to obtain, but determination of the SFC becomes more accurate at lower  $\psi_L$ . By contrast, the SMC provides reliable information on the water retention properties of the soil at high  $\psi_L$ . As soil dries, however, conventional SMC determination becomes time consuming and inaccurate. We therefore propose that water retention properties of soils are best obtained from conventional draining at higher matric potentials and from freezing at lower matric potentials.

## REFERENCES

- Anderson, D.W., and N.R. Morgenstern. 1973. Physics, chemistry, and mechanics of frozen ground: A review. p. 257-288. *In* Second Int. Conf. on Permafrost. Yakutsk, Siberia. 13-28 July 1973. Nat. Acad. of Sci., Washington, DC.
- Baker, J.M., and R.R. Allmaras. 1990. System for automating and multiplexing soil moisture measurement by time-domain reflectometry. *Soil Sci. Soc. Am. J.* 54:1-6.
- Baker, J.M., and R.J. Lascano. 1989. The spatial sensitivity of time-domain reflectometry. *Soil Sci.* 147:378-384.
- Baker, J.M., and E.J.A. Spaans. 1994. Ponded infiltration into frozen soil. p. 244. *In* Agronomy abstracts. ASA, Madison, WI.
- Banin, A., and D.M. Anderson. 1974. Effects of salt concentration changes during freezing on the unfrozen water content of porous materials. *Water Resour. Res.* 10:124-128.
- Black, P.B., and A.R. Tice. 1989. Comparison of soil freezing curve and soil water curve data for Windsor sandy loam. *Water Resour. Res.* 25:2205-2210.
- Campbell, G.S. 1988. Soil water potential measurement: An overview. *Irrig. Sci.* 9:265-273.
- Cary, J.W., R.I. Papendick, and G.S. Campbell. 1979. Water and salt movement in unsaturated frozen soil: Principles and field observations. *Soil Sci. Soc. Am. J.* 43:3-8.
- Colbeck, S.C. 1982. Configuration of ice in frozen media. *Soil Sci.* 133:116-123.
- Dirksen, C., and R.D. Miller. 1966. Closed-system freezing of unsaturated soil. *Soil Sci. Soc. Am. Proc.* 30:168-173.
- Klute, A. 1986. Water retention: Laboratory methods. p. 635-660. *In* A. Klute (ed.) *Methods of soil analysis. Part 1.* 2nd ed. Agron. Monogr. 9. ASA and SSSA, Madison, WI.
- Knight, J.H. 1992. Sensitivity of time domain reflectometry measurements to lateral variations in soil water content. *Water Resour. Res.* 28:2345-2352.
- Koopmans, R.W.R., and R.D. Miller. 1966. Soil freezing and soil water characteristic curves. *Soil Sci. Soc. Am. Proc.* 30:680-685.
- Lewis, G.N., and M. Randall. 1961. *Thermodynamics.* 2nd ed. McGraw-Hill, New York.
- List, R.J. 1951. *Smithsonian meteorological tables.* 6th ed. Smithsonian Inst., Washington, DC.
- Madsen, H.B., C.R. Jensen, and T. Boysen. 1986. A comparison of the thermocouple psychrometer and the pressure plate methods for determination of soil water characteristic curves. *J. Soil Sci.* 37:357-362.
- Miller, E.E., and R.D. Miller. 1956. Physical theory for capillary flow phenomena. *J. Appl. Phys.* 27:324-332.
- Miller, R.D. 1966. Phase equilibria and soil freezing. p. 193-197. *In* Int. Conf. Proc. Permafrost. Lafayette, IN. 1963. Natl. Acad. Sci.-Natl. Research Council, Washington, DC.
- Miller, R.D. 1972. Freezing and heaving of saturated and unsaturated soils. *Highway Research Record* 393:1-11.
- Miller, R.D. 1973. Soil freezing in relation to pore water pressure and temperature. p. 344-352. *In* Second Int. Conf. on Permafrost. Yakutsk, Siberia. 13-28 July 1973. Natl. Acad. Sci., Washington, DC.
- Miller, R.D. 1980. Freezing phenomena in soils. p. 254-299. *In* D. Hillel (ed.) *Applications of soil physics.* Academic Press, New York.
- Oliphant, J.L. 1985. A model for dielectric constants of frozen soils. p. 46-56. *In* D.W. Anderson and P.J. Williams (ed.) *Freezing and thawing of soil-water systems: A state of the practice report.* Tech. Council on Cold Regions Engineering/Am. Soc. of Civil Engineers Monograph. ASCE, New York.
- Patterson, D.E., and M.W. Smith. 1981. The measurement of unfrozen water content by time domain reflectometry: Results from laboratory tests. *Can. Geotech. J.* 18:131-144.
- Rhoades, J.D. 1986. Soluble salts. p. 167-179. *In* A.L. Page et al. (ed.) *Methods of soil analysis. Part 2.* 2nd ed. Agron. Monogr. 9. ASA and SSSA, Madison, WI.
- Schofield, R.K. 1935. The pF of water in the soil. p. 37-48. *In* E.M. Crowther et al. (ed.) *Trans. Int. Congr. Soil Sci.* Vol. 2. Oxford, UK. 30 July-7 Aug. 1935. Thomas Murby & Co., London.
- Spaans, E.J.A. 1994. The soil freezing characteristic: Its measurement and similarity to the soil moisture characteristic. Ph.D. diss. Univ. of Minnesota, St. Paul (Diss. Abstr. 95-01135).
- Spaans, E.J.A., and J.M. Baker. 1993. Simple baluns in parallel probes for time domain reflectometry. *Soil Sci. Soc. Am. J.* 57:668-673.
- Spaans, E.J.A., and J.M. Baker. 1995. Examining the use of TDR for measuring liquid water content in frozen soil. *Water Resour. Res.* 31:2917-2925.
- Stein, J., and D.L. Kane. 1983. Monitoring the unfrozen water content of soil and snow using time domain reflectometry. *Water Resour. Res.* 19:1573-1584.
- Tice, A.R., J.L. Oliphant, Y. Nakano, and T.F. Jenkins. 1982. Relationship between the ice and unfrozen water phases in frozen soil as determined by pulsed nuclear magnetic resonance and physical desorption data. Report 82-15. U.S. Army Cold Regions Research and Engineering Lab., Hanover, NH.
- Williams, P.J. 1964. Unfrozen water content of frozen soils and soil moisture suction. *Geotechnique* 14:231-246.